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4 Interannual to Decadal Variability of Atlantic Water in the Nordic and
5 Adjacent Seas
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46 **Abstract**

47 Warm salty Atlantic Water is the main source water for the Arctic Ocean and thus plays
48 an important role in the mass and heat budget of the Arctic. This study explores
49 interannual to decadal variability of Atlantic Water properties in the Nordic Seas area
50 where Atlantic Water enters the Arctic, based on a reexamination of the historical
51 hydrographic record for the years 1950-2009, obtained by combining multiple data sets.
52 The analysis shows a succession of four multi-year warm events where temperature
53 anomalies at 100m depth exceed 0.4°C, and three cold events. Three of the four warm
54 events lasted 3-4 years, while the fourth began in 1999 and persists at least through 2009.
55 This most recent warm event is anomalous in other ways as well, being the strongest,
56 having the broadest geographic extent, being surface-intensified, and occurring under
57 exceptional meteorological conditions. Three of the four warm events were accompanied
58 by elevated salinities consistent with enhanced ocean transport into the Nordic Seas, with
59 the exception of the event spanning July 1989-July 1993. Of the three cold events, two
60 lasted for four years, while the third lasted for nearly 14 years. Two of the three cold
61 events are associated with reduced salinities, but the cold event of the 1960s had elevated
62 salinities. The relationship of these events to meteorological conditions is examined.
63 The results show that local surface heat flux variations act in some cases to reinforce the
64 anomalies, but are too weak to be the sole cause.

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67

68 **1. Introduction**

69 The recent dramatic warming of Arctic Ocean surface temperatures and shrinkage of sea
70 ice coverage have been accompanied by a warming and salinification of the Atlantic
71 Water Mass flowing into the Nordic Seas from the Atlantic (*Holliday et al., 2008;*
Piechura and Walczowski, 2009; Matishov et al., 2009; Dmitrenko et al., 2008, 2009).

73 Warmings of this water mass have also occurred in the 1920s-1930s, 1960s, 1970s, and
74 1990s (*Swift et al., 2005; Levitus et al., 2009a; Matishov et al, 2009*) raising the question
75 of how different the recent warming is from past warmings. As a contribution to
76 addressing this question we revisit the interannual-to-decadal variability of the Atlantic
77 Water mass in the Nordic Seas during the past 60 years and its relationship to
78 hemispheric climate variability using a single combined hydrographic data set that builds
79 on the newest release of the World Ocean Database (*Boyer et al., 2009*).

80

81 The Nordic Seas is the region north of the Greenland-Scotland Ridge and south of the
82 Arctic Ocean, including the fairly shallow Barents, and deep Norwegian, Iceland, and
83 Greenland Seas (*Hansen and Osterhus, 2000*) (**Fig. 1**). The inflow of the Atlantic Water,
84 which is characterized by warm temperatures and high salinity, occurs via three main
85 branches: approximately 1Sv ($=10^6 \text{ m}^3 \text{s}^{-1}$) passes west of Iceland in the North Icelandic
86 Irminger Current, approximately 3.3-3.5Sv travels over the Iceland Faroe Ridge in the
87 Faroe Current and approximately 3.7Sv passes in the form of inflow through the Faroe
88 Shetland Channel (see *Hansen and Osterhus, 2000; Hansen et al, 2003* and references
89 therein). The two western branches are mainly fed by water from the North Atlantic

90 Current while inflow through the Faroe Shetland Channel is mainly the warmer and more
91 saline Eastern North Atlantic Water (e.g. *Hakkinen and Rhines, 2009; Hakkinen et al.,*
92 *2011a*). Flow in the eastern branches merge towards the Norwegian coast in the
93 Norwegian Sea, but retain two distinct cores (*Mork and Blindheim, 2000; Orvik and*
94 *Niiler, 2002; Holliday et al. 2008*).

95

96 This flow branches with one branch of approximately 1.1 to 1.7Sv travelling eastward
97 into the Barents Sea, and the other flowing northward in the West Spitsbergen Current
98 through Fram Strait (*Ingvaldsen et al., 2004; Skagseth et al., 2008*). In the vertical it
99 extends from a maximum of 600m depth to the base of the Arctic Surface Water. As this
100 Atlantic Water is transported poleward its properties are altered through mixing so that
101 temperature and salinity classes vary significantly by location. In the shallow Barents
102 Sea cooling and freshening leads to an approximately linear relationship between
103 temperature and salinity ranging from 8°C and 35.13 psu down to 0°C and 34.87psu or
104 colder (*Schauer et al., 2002; Smolyar and Adrov, 2003*). At Fram Strait *Schauer et al.*
105 (*2004*) define a minimum temperature for Atlantic Water (>1°C), *Schlüchtholz and*
106 *Gosczko (2006)* define both temperature and salinity minimums (>2°C and >34.9psu),
107 *Saloranta and Haugan (2001)* and *Piechura and Walczowski (2009)* make similar
108 choices (>0°C, >34.93psu), while *Marnela et al. (2008)* define a density range
109 ($\sigma_{\theta} = 27.70$ to 27.97). For this study we define the Atlantic Water Zone (AWZ) as the
110 geographic region circumscribed by the time mean 35psu contour at 100m depth north of
111 63°N, (see **Fig.1**). The southern limit of this domain is chosen to exclude most of the
112 ocean south of the Iceland-Faroe Ridge from these calculations. Calculations are done

113 over a depth range of 0/600m which spans the depth range of the Atlantic Water in this
114 region. The data coverage is insufficient to allow more sophisticated isopycnal or
115 isohaline analyses.

116

117 Superimposed on seasonal variations in transports and stratification (e.g., *Ingvaldsen et*
118 *al., 2004*), Atlantic Water properties undergo strong interannual to decadal variations.

119 One of the best continuous records is available at Ocean Weather Station M, maintained
120 in the Norwegian Sea (66°N, 2°E) since 1948. The Station M record shows a warming
121 trend at the surface and a decadal succession of warmings at the surface and in the 50-
122 200m layer in the early 1960s, and the early to mid-1970s, (*Blindheim et al 2000;*
123 *Polyakov et al., 2004*). The transition from warm, salty conditions of the early to mid-

124 1970s to cooler and fresher conditions of the Great Salinity Anomaly appears
125 dramatically in this region during 1976-1981 (*Dickson et al., 1988*). Further discussion
126 of this remarkable event is available in *Belkin et al. (1998)* and *Belkin (2004)*. Warm
127 conditions returned in early 1990s followed by cooling in the mid-1990s. Many of these
128 warmings are associated with enhanced salinity superimposed on a multi-decadal

129 freshening trend. The Kola Bay meridian transect (lying along the 33°E meridian) offers
130 another long time series record of conditions within the Barents Sea extending back to the
131 beginning of the 20th century and, among other features, delineating the 1920s-1930s
132 warming (*Matishov et al, 2009*). Since 1960 the time series suggests that along this
133 meridian there were also a set of warm events in the 1960s, early to mid-1970s, 1983-4,
134 and 1989-1994, and the 2000s with prominent cold events in the late-1960s, late-1970s,
135 and late-1990s. *Saloranta and Haugan (2001)* and *Polyakov et al (2004)* describe a time

136 series for the region northwest of Svalbard that also shows variability on similar time
137 scales.

138

139 To complement these rare time series observations, a number of studies have examined
140 conditions during specific years. The early 1960s event appears at coarse resolution in
141 the Environmental Working Group survey described in *Swift et al. (2005)*. Interestingly,
142 the modeling study of *Gerdes et al. (2003)* suggests this event to be the result of the
143 impact of a melt-induced halocline on net surface heat loss.

144

145 *Furevik (2000; 2001)* and *Saloranta and Haugan (2001)* track spatial aspects of heat
146 anomalies for the warm and cold events of the 1980s and 1990s. These studies make the
147 case for the importance of ocean advection in giving rise to the 1983-1984 warm event
148 and the 1986-1988 cold event, but argue that the 1990-1992 warm event may be most
149 closely related to changes in surface flux. *Furevik (2000)* also introduces the use of
150 observations of SST to track subsurface temperature anomalies.

151

152 Studies of the warming which has developed since 1999 have shown this latter to be
153 dramatic (certainly the strongest warming since the 1930s) and also to have
154 characteristics, such as its vertical structure, which set it apart from previous warmings
155 (*Matishov et al., 2009*). *Walczowski and Piechura (2006)*, and *Dmitrenko et al. (2008)*
156 identify indications of anomalous temperature advection through Fram Straits and also
157 through the Barents Sea into the northern Laptev Sea. Based on observations at Fram
158 Strait *Schauer et al. (2004)* ascribe the advective cause of the warming to both increasing

159 temperatures of the source waters and to increases in volume transport rates. Consistent
160 with this, *Hakkinen and Rhines (2009)* examine surface currents in the source waters of
161 the subpolar gyre and identify the presence of anomalous poleward advection of warm
162 and saline waters.

163

164 The processes giving rise to these warm and cold events have been examined in many
165 modeling studies (*Zhang et al. 1998; Karcher et al., 2003; Hu et al., 2004; Mauritzen, et*
166 *al., 2006; Sando and Furevik, 2008; and Semenov et al., 2009*). The models with
167 specified meteorology show significant ability to reproduce past anomalies in the Nordic
168 Seas. Analysis of the results generally highlights the importance of changes in rate of
169 advection of Atlantic Water entering the Nordic Seas.

170

171 This conclusion regarding the importance of ocean advection has also been used to
172 explain an observed relationship between the December-February North Atlantic
173 Oscillation (NAO) Index of winter sea level pressure and the temperature anomalies in
174 the Nordic Seas (*Swift et al., 1998; Grotfendt et al., 1998; Dickson et al., 2000; Mork*
175 *and Blindheim, 2000; Saloranta and Haugan, 2001; Polyakov et al., 2004; Furevik and*
176 *Nilsen, 2005; and Schlichtholz and Gosczko, 2006*). However, the recent warming
177 which has occurred under fairly neutral NAO conditions is an example of a situation
178 when the relationship breaks down. *Bengtsson et al. (2004)* and *Sando and Furevik*
179 *(2008)* propose that the key meteorological parameter is the local winds in the narrow
180 zone between Spitsbergen and Norway. They argue, following *Zhang (1998)*, that it is
181 only these local winds which drive Atlantic Water transport into the Barents Sea thus

182 increasing the spread of Atlantic Water (see also *Dmitrenko et al., 2009*). *Häkkinen et al.*
183 (*2011*) show evidence to suggest that winter meteorological conditions over the
184 subtropical gyre are an additional factor contributing to warming by forcing anomalous
185 quantities of Atlantic Water into the Nordic Seas. This basin-scale contribution reached a
186 peak around 1990 suggesting that this effect may have been important for the early 1990s
187 warming. In a coupled modeling study *Semenov et al. (2009)* suggest that sea ice and
188 surface heat flux changes also play a role in a feedback cycle associated with these
189 transport variations.

190

191 This study is an effort to reexamine the spatial and temporal structure of the anomalies of
192 Atlantic Water in the Nordic Seas with a combined hydrographic data set and then use
193 the results to explore their relationship to atmospheric variables for the years since 1950.

194

195 **2. Methods**

196 This study relies foremost on the set of 420,199 temperature profiles and 258,912 salinity
197 profiles (60°N-90°N, 50°W-80°E) in the National Oceanographic Data Center's World
198 Ocean Database 2009 (WOD09, *Boyer et al., 2009*), including all instrument types which
199 measure temperature or salinity, for the sixty-year period 1950-2009. The version used
200 here is the standard level data which includes quality control carried out by the National
201 Oceanographic Data Center. We have also applied a secondary quality check (including:
202 checks for static stability, profile depth, buddy-check, and comparison to climatological
203 profiles) which eliminates 3% of the salinity profiles and 1.8% of the temperature
204 profiles. In addition to the WOD data we include all observations from the International

205 Council of the Exploration of the Seas database (approximately 50,000 unique profiles),
206 the Woods Hole Oceanographic Institution Ice-Tethered Profile data, the North Pole
207 Environmental Observatory data, and the Nansen and Amundsen Basin Observational
208 System data (just a handful in our domain).

209

210 Many of the additional profiles were duplicates, but after eliminating these the additions
211 increase our basic temperature data set by 58,469 profiles to a total of 478,668
212 temperature profiles, and the basic salinity data set by 52,314 profiles to 311,226 profiles,
213 with the greatest increases in the northern North Atlantic and Barents Sea. The data set
214 does contain some Expendable Bathythermograph temperature profiles, which are known
215 to contain a time-dependent warm bias, notably during the 1970s. WOD09 contains a
216 correction for this bias based on the work of *Levitus et al. (2009b)* which seems to
217 remove it as an area of concern for this study.

218

219 The resulting observation distribution shows increases in coverage in the 1950s due to the
220 introduction of drifting stations as well as aircraft and submarine surveys, and again in
221 the 1980s due to a succession of scientific experiments (**Fig. 2**). Seasonally, the number
222 of temperature observations peaks in summer at nearly 800 per month reducing to only
223 250 per month in winter.

224

225 All temperature and salinity profiles that are not already on standard levels are first
226 interpolated to National Oceanographic Data Center standard levels using linear
227 interpolation. After a simple set of quality controls the data are binned into

228 $1^{\circ} \times 1^{\circ} \times 1$ month bins. To check the data set we have computed monthly climatologies of
229 temperature and salinity and compared them to the monthly climatology to the Polar
230 science center Hydrographic Climatology PHC3.0 (*Steele et al., 2001*). This new
231 climatology is generally consistent with PHC3.0, but with some persistent patterns of
232 $\pm 0.5^{\circ}\text{C}$ temperature and $\pm 0.1\text{psu}$ salinity differences on 100km scales in the upper 200m.
233 We believe these differences result from the use in the current study of a more extensive
234 data set with a more recent time-centering than PHC3.0 (which was based only on data
235 collected prior to 1998). A simple *Cressman (1959)* gridding scheme is used to smooth
236 the results in order to present the results graphically. Anomalies of AWZ heat content are
237 evaluated by computing the volume integral of temperature times its heat capacity in the
238 upper 600m of the AWZ, removing the time mean and smoothing with a two-year filter.
239 Later we will identify warm and cold events as those that exceed $\pm 2 \times 10^{20}\text{J}$.
240
241 Because of persistent limitations of the spatial coverage of the hydrographic data set we
242 complement our examination of hydrography by also examining the monthly
243 NOAA/National Climate Data Center OI SST v2. The OI SST data set is based on
244 satellite infrared emissivity and is available at $0.5^{\circ} \times 0.5^{\circ}$ resolution daily for the period
245 January 1982 to December 2010 (*Reynolds et al., 2007*). While its spatial coverage is
246 excellent OI SST is subject to potential sources of bias including: undetected stratus
247 clouds and aerosol effects, unresolved diurnal effects, as well as error in interpretation of
248 OI SST as if it were a measurement of bulk SST. These errors are particularly a concern
249 during the years 1995-2002 due to problems with the satellite instruments (*Podesta et al.,*
250 *2003*).

251

252 To evaluate the bias in OI SST, monthly binned temperature observations at 10m depth
253 (deep enough to eliminate diurnal effects) throughout the domain 60°N-90°N, 50°W-80°E
254 were matched with collocated monthly average OI SST observations for the years 1982-
255 2009. Examination of these temperature differences reveals a temperature-dependent
256 systematic 1°C warm bias in OI SSTs for SSTs cooler than 2°C becoming a 0.5-1.0°C
257 cold bias for SSTs above 4°C. After removal of this bias through a simple geographical,
258 and temperature-independent correction, the recomputed analysis of collocated monthly
259 SST differences shows a ±2°C essentially random difference suggesting that this
260 corrected monthly average OI SST should provide an unbiased estimate of bulk SST with
261 an accuracy in the range of previous estimates (e.g. *Key et al., 1997*).

262

263 Finally, when examining the causes of anomalous rate of heat storage in the Nordic Seas
264 we compare to four estimates of net surface heat flux. Three are based on atmospheric
265 reanalyses: the National Centers for Environmental Prediction/National Center for
266 Atmospheric Research (NCEP/NCAR) reanalysis of *Kalnay et al. (1996)*, which covers
267 the full period of this study; the European Centers for Medium Range Weather Forecasts
268 ERA-40 analysis (*Uppala et al, 2005*), which covers the period 1958-2001; and the
269 updated ERA-Interim, which spans the period 1989-2009 (*Dee and Uppala, 2009*). The
270 fourth we consider is the Woods Hole Oceanographic Institution OAFlux/ISCCP analysis
271 of *Yu et al. (2008)*, which is based on a combination of other products, weighted to match
272 observations, and is available for the period 1983-2007. The first three also provide
273 individual radiative and thermodynamic flux components.

274

275 **3. Results**

276 The time-mean horizontal structure of temperature and salinity at 100m shows a strong
277 positive relationship between temperature and salinity due to the presence of an intrusion
278 of warm salty Atlantic water and its gradual dilution. Interestingly, and helpfully for this
279 study, the geographic location of this Atlantic Water is readily identified by the area
280 defined by the 35psu isohaline (the AWZ), which remains generally stable with time.

281 The shape of the AWZ reflects the fact that there is poleward path into the Nordic Seas
282 (A-B on **Fig. 1**) which then branches with the northward branch passing through Fram
283 Strait (B-C) and an eastward branch extending into the Barents Sea (B-D). The eastward
284 branch shows more rapid cooling with distance than the northward branch and significant
285 freshening as well.

286

287 The vertical structure of temperature and salinity along path A-B shows a subsurface
288 salinity maximum with a core depth of 100-200m and a layer thickness that increases
289 from less than 400m at point A to nearly 600m by point B, now generally overlain by
290 cold, fresh surface water (**Fig. 3**). Interestingly, examination of the spatial structure of
291 this overlying fresh surface layer suggests that it thins or disappears in a narrow band
292 overlying the central core of the Atlantic Water between A-B. Along this path the high
293 salinity (>35psu) layer approximately corresponds to a density range of 27.25-27.85 kg
294 m⁻³ and a temperature range of 2°C-9°C. Along the northward branch (B-C) densities
295 increase by approximately 0.25 kg m⁻³ while temperatures cool to below 5°C. Along the
296 eastward branch (B-D) density also increases by 0.25 kg m⁻³, but the temperature cools

297 even more and salinities drop below 35psu. The complexity of these water mass changes
298 explains the multiple regionally-dependent definitions of Atlantic Water described in the
299 *Introduction*. Seasonal changes are most pronounced in the surface layer (**Fig. 1** insets).

300

301 We next examine the variability of temperature and salinity anomalies about their
302 monthly climatology, averaged spatially throughout the AWZ, and annually averaged
303 with a 12-month running filter in **Fig. 4** to eliminate seasonal variability and increase
304 statistical confidence. The relatively shallow Barents Sea means that the horizontal
305 extent of the integrals vary with depth. The record reveals that since 1950 there have
306 been four distinct warm events spanning multiple years: July 1959-July 1962, October
307 1971- August 1975, July 1989-July 1993, and March 1999- December 2009 and beyond,
308 which we label W1-W4 (emulating the notation of *Furevik, 2001*). In-between these we
309 find three cold events: April 1965- February 1969, July 1976- January 1989, and August
310 1994- March 1998 which we label C1-C3. As discussed in the *Introduction* each of these
311 events have also been identified in other observational studies.

312

313 W1 (1959-1962) first appears as a 0.3°C nearsurface warm temperature and 0.04 psu high
314 salinity anomaly in late 1959 whose density effects nearly compensate (**Fig. 4**). The
315 temperature anomaly rapidly deepens to 500m and then more gradually to 1000m
316 (well below the layer containing the Atlantic Water core) by 1964, accompanied by a
317 similar deepening of the salinity anomaly. Examination of the spatial structure of the
318 anomaly (**Fig. 5**) shows the temperature and salinity anomalies to be most
319 pronounced in the south, extending into the Barents Sea, beyond the AWZ towards

320 Novaya Zemlya. Lack of data coverage limits our ability to see the northward
321 extension of this event.

322 C1 (1965-1969) is a -0.3°C cold anomaly that persists until early 1969 when salinities
323 increase throughout the upper 800m by more than 0.04psu. As a result of this cooling
324 and salinification, stratification in the upper ocean above 400m weakens significantly
325 throughout the late-1960s and early 1970s and potential density increases by 0.04kg
326 m⁻³. An examination of the spatial structure of this event shows the maximum cold
327 anomaly to be in the Norwegian Sea south of Svalbard while the salinity anomaly is
328 displaced further southward (**Fig. 6**). The eastward extent of either is unknown due to
329 limitations in data coverage during this decade.

330 W2 (1971-1975) has a deeper structure than W1 with a maximum temperature anomaly
331 between 600-700m depth and a correspondingly deep, but weak, positive salinity
332 anomaly (indeed, the salinity anomaly precedes this warm event). Geographically,
333 the deep temperature anomaly primarily occurs east of Iceland with shallower
334 temperature anomalies extending into the Barents Sea, while the positive salinity
335 anomaly is most pronounced in the Barents Sea at a time when the subpolar gyre is
336 anomalously fresh (**Fig. 5**).

337 C2 (1976-1989) is the longest lasting anomaly considered here (13 years), but the major
338 cold anomaly occurs during the first four years, a part of the event which has been
339 associated in the literature (e.g. *Dickson et al., 1988; Belkin et al., 1998*) with the
340 return of the Great Salinity Anomaly to the Nordic Seas. During these early years
341 temperatures between 0/300m decrease by more than -0.45°C and salinities decrease
342 by more than -0.08psu. As in the case of W1 these temperature and salinity

343 anomalies nearly compensate so the impact on density and thus stratification is weak.
344 This cold/fresh anomaly has broad spatial structure, spanning much of our domain
345 (**Fig. 6**). During the later years of C2 temperatures in the upper 500m remain
346 anomalously cool, but with only weak salinity anomalies so that surface density
347 increases. C2 represents a transition point in our record in that the water column is
348 significantly more stable in the years following 1989 than in prior years due to the
349 persistence of the freshening which began in the late-1970s.

350 W3 (1989-1993) is a five-year long 0.3°C warming of the upper 500m, but as there is no
351 compensating salinity anomaly its effect on increasing the stratification of the base of
352 the Atlantic Water was substantial. This warm anomaly is most pronounced west of
353 the Barents Sea Opening and throughout the Barents Sea at a time when the subpolar
354 gyre is anomalously cool (**Fig. 5**). In contrast to the other warm events W3 is
355 associated with reduced salinities throughout much of the domain except along the
356 southern Norwegian coast.

357 C3 (1994-1998) is a short-lived but 700m deep weak cold/fresh event which is density
358 compensated. The spatial structure of this anomaly is not well-defined due to
359 limitations of the hydrographic sampling during the 1990s, however the anomaly
360 appears to extend at least partway into the Barents Sea (**Fig. 6**).

361 W4 (1999-2009+) is a strong, surface intensified warm event with temperature anomalies
362 in the upper 300m that exceed 0.45°C . This warming occurs throughout the Nordic
363 Seas except Denmark Strait and is accompanied by a similarly broad, but weak
364 salinity anomaly (**Fig. 5**). As a result W4 is associated with significantly enhanced
365 stratification with surface density decreasing by 0.045kg m^{-3} .

366

367 The surface intensification of the temperature anomaly associated with W4 (e.g. **Fig.**
368 **4)** allows us to use SST to examine its basin-scale and detailed temporal structure
369 (**Fig. 7**). SST shows that the warming spans the entire northern North Atlantic sector
370 including such marginal seas as the North Sea and the Baltic, but is more pronounced
371 in the western side of the Atlantic basin than the central/east. The SST also shows
372 that warming is present in both winter and summer, although it is less pronounced in
373 marginal seas in winter. In contrast, in the Barents Sea the warm anomaly is less
374 evident in summer, perhaps due to capping by low salinity Surface Water. Averaged
375 over the extended domain (60°W - 80°E , 40°N - 90°N , excluding marginal seas) the
376 anomaly was quite prominent in the summers of 2001 and 2003, but then remains
377 quite warm both winter and summer from 2005-2008. In 2009 it is reduced, but in
378 2010 the SST anomaly returns in intensified form.

379

380 As discussed in the *Introduction* both advective effects and variations in surface heating
381 have been suggested as causes of temperature anomalies in the Nordic Seas. We explore
382 the potential contribution of anomalies of surface heat flux over the AWZ itself by first
383 computing anomalous AWZ heat content in the upper 600m and then differentiating this
384 quantity to obtain rate of heat storage (**Fig. 8**). The time series of 0/600m heat content
385 shows each of the warm and cold events identified above exceed $\pm 2 \times 10^{20}\text{J}$ (equivalent
386 to¹ a 0/600m average temperature anomaly of $\pm 0.12^{\circ}\text{C}$). The record maximum heat
387 content anomaly of $> 1 \times 10^{21}\text{J}$ occurs at the peak of warming associated with W4 in early

¹ A reviewer notes that the observations of Piechura and Walczowski (2009) are in the West Spitsbergen Current which typically lags the southern Norwegian Sea and Faroe Shetland Channel by 1-4 years.

388 2005 (a year before the peak appears in the AREX cruise data of *Piechura and*
389 *Walczowski, 2009* and more than a year before the peak in SST) while the minimum of <
390 $-9 \times 10^{20} \text{ J}$ occurs during C2 in 1979. Interannual variations in rate of heat storage fall in
391 the range of $\pm 20 \text{ Wm}^{-2}$.

392

393 It is evident from comparison of the time series (**Fig. 8** upper panel) that the NAO Index
394 (either annual or DJF) is an uneven predictor of warming of the AWZ. W1, W2 and W3
395 are all associated with positive values of the NAO Index as well as positive wind stress
396 curl anomalies over the Atlantic Water domain (although we note that W3 is not
397 associated with high salinities). Cold event C1 is associated with a negative NAO Index
398 anomalies, but not with low salinities, while C2 and C3 do not seem to be associated with
399 extrema of the NAO Index. We anticipate that the impact of NAO on the AWZ is
400 strongly modulated by local forcing associated with the appearance of anomalous
401 westerly winds in the Barents Sea, reminiscent of the suggestion of local forcing by
402 *Bengtsson et al. (2004)* and *Sando and Furevik (2008)*.

403

404 Finally, we consider the direct effect of meteorological conditions on net surface heat
405 flux by comparing anomalous rate of heat storage in the AWZ to anomalous net surface
406 heat flux averaged over the AWZ. Four anomalous flux estimates considered here differ
407 from each other by a substantial $\pm 5\text{-}10 \text{ Wm}^{-2}$. However, some common features are
408 evident. Of the two surface flux estimates that span multiple decades NCEP/NCAR has
409 significantly larger anomalies than ERA-40. Examination of the individual flux
410 components in both data sets shows an increased sensible heat flux of up to 10 Wm^{-2} due

411 to warm air advection. Despite these differences in amplitude, the phases of the
412 anomalies in the two products are in reasonable agreement with each other and both are
413 too small by a factor of 2-3 to explain the heat storage anomalies. We conclude that heat
414 flux offers a weak positive contribution to heat storage (with heat flux anomalies lagging
415 heat storage anomalies somewhat).

416

417 A similar conclusion can be drawn from the combined set of four flux estimates for the
418 more recent two decades when all are available. Around 1990, for example, the surface
419 flux estimates show that net downward surface heat flux increases for several years.

420 Examination of the individual flux components shows this increase is due to an increase
421 of sensible heat flux of up to 10Wm^{-2} due to warm air advection. But, this surface flux
422 anomaly is too small by a factor of two to explain the increase in heat storage, an
423 explanation which had been proposed by *Overland et al. (2008)*. Other factors such as
424 wind-induced changes in the rate of Atlantic Water entering the Nordic Seas (discussed
425 above and inferred in **Fig. 8** lower panel from the difference between storage and surface
426 flux) must be invoked to explain these heat storage anomalies.

427

428 **4. Summary**

429 Here we report on a reexamination of subseasonal anomalies of Atlantic Water in the
430 Nordic Seas area of the Arctic Mediterranean for the 60-year period 1950-2009 using a
431 combined set of hydrographic observations drawn from several data sets. The
432 reexamination reveals a succession of four warm and three cold events during this period
433 which exceed a threshold of $\pm 2 \times 10^{20}\text{J}$ in the Atlantic Water Zone for two years, W1: July

434 1959-July 1962, W2: October 1971- August 1975, W3: July 1989-July 1993, and W4:
435 March 1999- December 2009 and beyond, C1: April 1965- February 1969, C2: July
436 1976- January 1989, and C3: August 1994- March 1998. Many of these events show that
437 they extend eastward into the Barents and Kara Seas. Their extensions northward past
438 Fram Straits are uncertain due to lack of data coverage. The events have other features in
439 common. Warm events are generally characterized by elevated salinities (the exception
440 is W3, July 1989-July 1993), while two of three cold events are characterized by reduced
441 salinities. This positive relationship is consistent with the volumetric estimates *Levitus et*
442 *al. (2009a)* in the Barents Sea, suggesting the importance of anomalous advection of
443 Atlantic Water. Most of the events satisfying our definition persist for 3-4 years. The
444 first of two exceptions is C2 which lasted 13 years. C2 is unusual because the main
445 anomaly actually spans only the first four years and also because the event seems to have
446 permanently increased stability of the water column by lowering salinity of the upper
447 layer. The second exception is the recent and intense W4, spanning 1999 to the end of
448 our analysis in 2009.

449

450 A number of previous studies have also examined multi-year variability of the properties
451 of the Atlantic Water in this region, as reviewed in the *Introduction*. A notable study by
452 Furevik (2001) uses observations from five hydrographic sections to consider variability
453 during the 17-year period 1980-1996 relative to the mean of that period. Interestingly,
454 this period begins just after the most intense part of C2 and ends just before the onset of
455 the intense W4. One of the distinctions of these results from ours is that Furevik (2001)
456 highlights a warm event during 1983-4 (also evident in the Kola Bay transect) which

457 would fall in the middle of our second cold event C2. This 1983-4 warming is in fact
458 also evident in our results (e.g. **Fig. 4**) but is of lower amplitude than other events in our
459 record and thus does not meet our definition of a warm event.

460 In the second part of this paper we consider the relationship of the temperature anomalies
461 in the Nordic Seas to changing meteorological conditions. As discussed in the
462 *Introduction*, there is a (rather limited) relationship between these temperature anomalies
463 and the NAO Index, which in turn is associated with a northward displacement of storm
464 tracks as well as the strength of monthly mean westerlies in this region. Year to year
465 changes in synoptic wind variability within this region may turn out to be a key
466 additional variable (*Hakkinen et al. 2011b*). W4, which has occurred during a period of
467 near-neutral annual NAO Index, is an example of a case where the Index is a poor
468 predictor. A better predictor is the frequency of occurrence of wintertime atmospheric
469 blocking events. We also explore the possibility that the temperature anomalies in the
470 Nordic Seas may be explained by local anomalies of net surface heat flux, by comparing
471 heat storage in the Atlantic Water Zone with four alternative estimates of net surface flux.
472 We find that while the estimates are themselves not very consistent, they are all too small
473 by a factor of 2-3 to explain the observed heat storage anomalies.

474

475 Having ruled out anomalies of net surface heat flux, our results support the results of a
476 number of transport observation and modeling studies reviewed in the *Introduction* that
477 suggest anomalous advection of Atlantic Water into the Nordic Seas as the more
478 plausible explanation for the appearance of these anomalies (e.g. *Zhang et al., 1998*).

479 However, issues related to the underlying dynamics of this process and the potential role
480 of interactions with the overlying atmosphere remain.

481

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494

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- 650

651 Figure Legends

652

653 **Figure 1** Time mean temperature (contours) and salinity (colors) for the 60-year period
654 1950-2009 at 100m depth. The Nordic Seas Atlantic Water Zone (AWZ) used in
655 subsequent calculations is defined as the area circumscribed by the 35psu contour
656 north of 63°N. The Fram Strait (ABC) and Barents Sea (ABD) transects are
657 indicated as well. Upper left and lower right insets shows salinity (psu) and
658 temperature (°C) as a function of depth (m) averaged over the AWZ. Summer
659 (July-October) in red and winter (January-May) in blue.

660

661

662 **Figure 2.** (a) Number of temperature (black) and salinity (green) observations in the
663 sector 50°W-80°E, 60°N-90°N with time at 100m 1950-2009. Insets show
664 monthly average and vertical distribution of temperature and salinity
665 observations. (b) Spatial distribution of temperature (left) and salinity (right)
666 observations per 100km² square 1950-2009. Insets show monthly average and
667 vertical distribution of temperature and salinity observations.

668

669 **Figure 3.** Time mean temperature (solid contours), potential density (dashed contours),
670 and salinity (colors) for the 60-year period 1950-2009 computed in a 2° wide band
671 perpendicular to the track along the Fram Strait (ABC, upper panel) and Barents
672 Sea (ABD, lower panel) transects and plotted as a function of depth and distance
673 along transect in kilometers.

674 **Figure 4** Annual temperature, salinity, and potential density anomalies from their 60-year
675 mean in the AWZ as a function of depth and time. Four warm periods and three
676 cold periods are indicated. Seasonal profiles of temperature and salinity are
677 provided in **Fig. 1**.

678

679 **Figure 5** Geographic structures of temperature and salinity anomalies relative to the
680 climatological monthly cycle at 100m depth for the four warm events (shading).
681 Contours show the AWZ for the individual events (bold) and the mean position of
682 the AWZ (thin). Regions with insufficient data sampling are shaded light grey.

683

684 **Figure 6** Geographic structures of temperature and salinity anomalies relative to the
685 climatological monthly cycle at 100m depth for the three cold events (shading).
686 Contours show the AWZ for the individual events (bold) and the mean position of
687 the AWZ (thin). Regions with insufficient data sampling are shaded light grey.

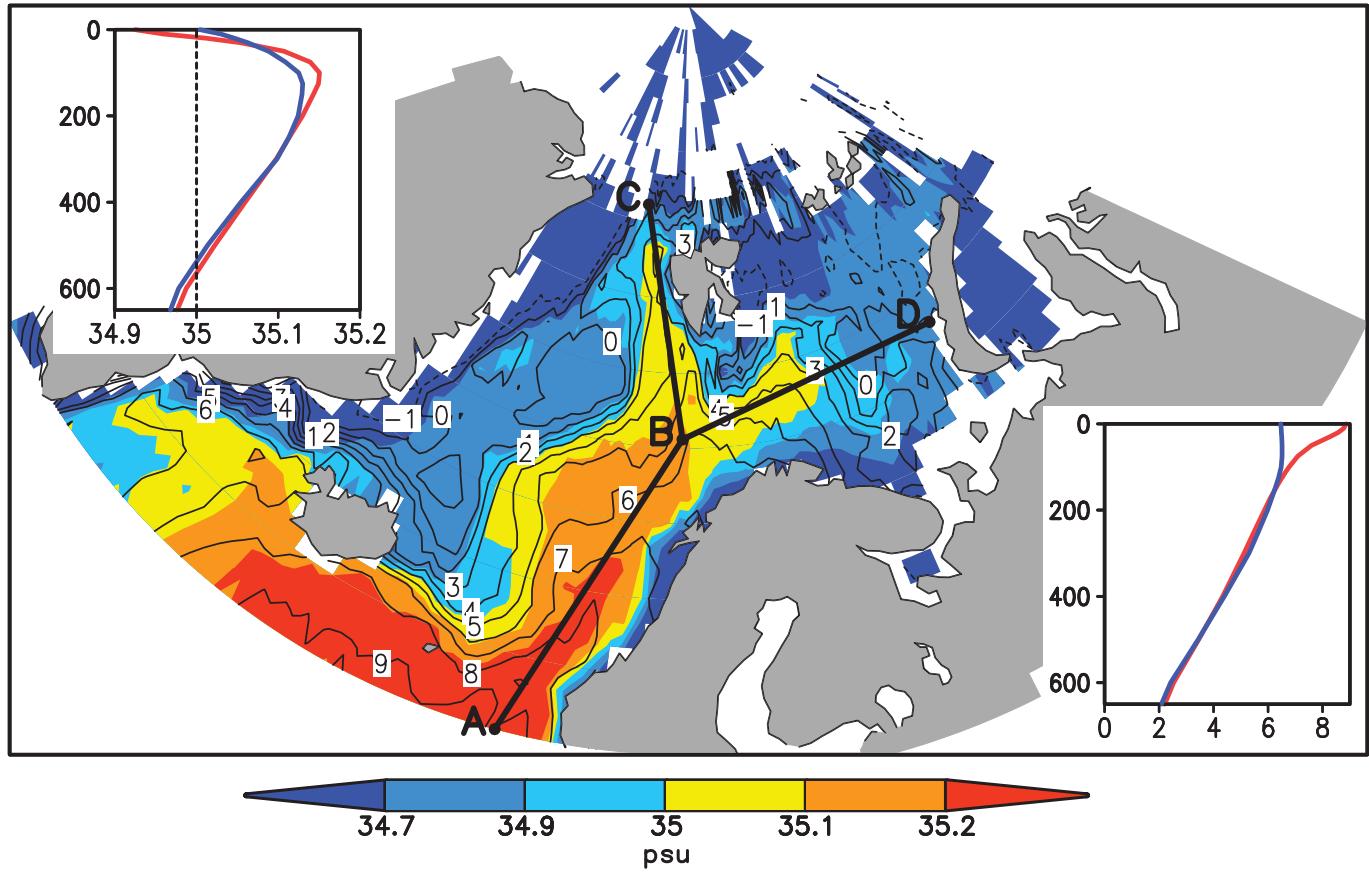
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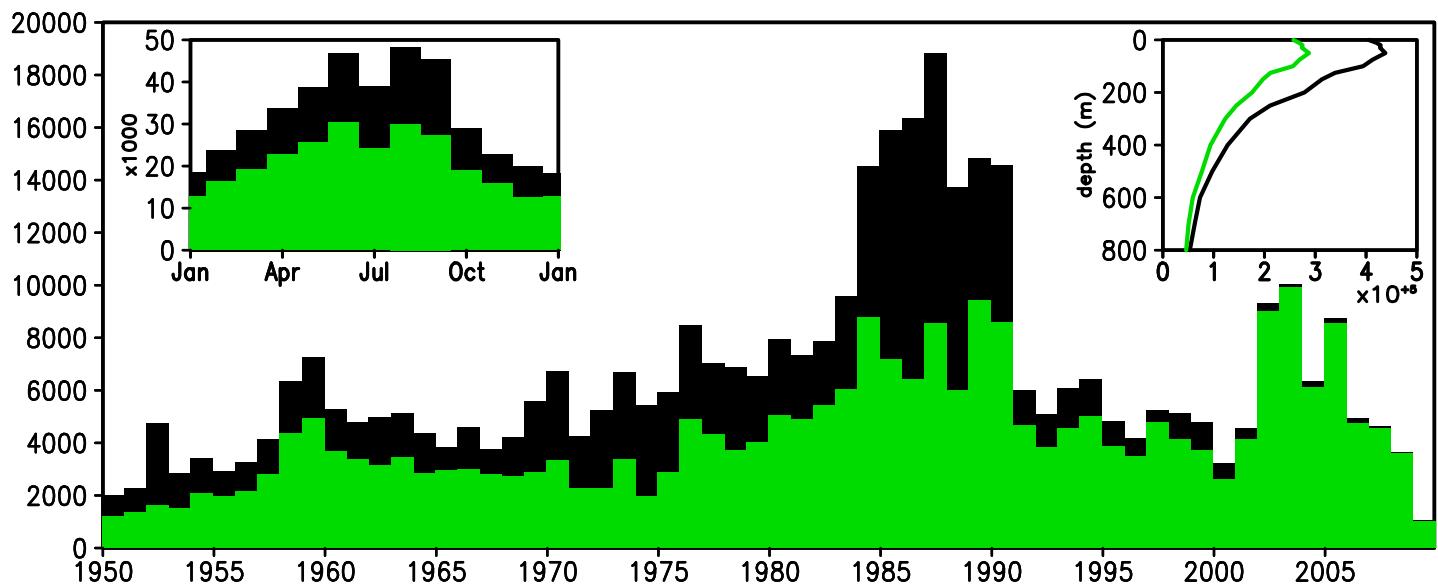
689 **Figure 7**. Upper and middle panels show SST anomaly for summer (May-Oct) and
690 winter (Nov-Apr) during W4 (March 1999 - December 2009) relative to the
691 climatological monthly cycle 1982-2009. Lowest panel shows time series of
692 monthly anomalies averaged over the extended domain 60°W-80°E, 40°N-90°N
693 (excluding marginal seas). Colors highlight (red) summer and (blue) winter
694 seasons.

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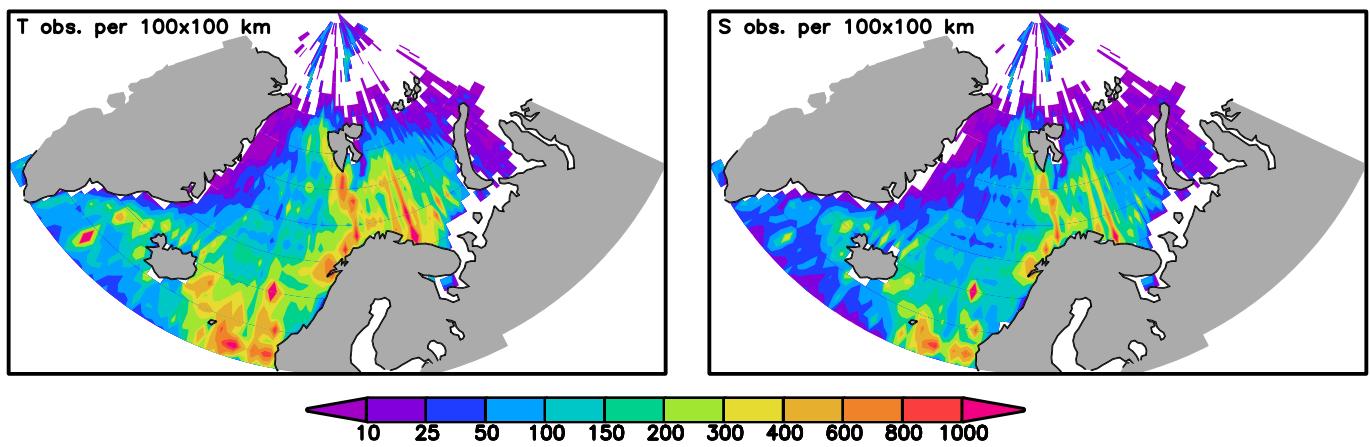
$$HC = \rho C_p \int_{600}^0 T' dz$$

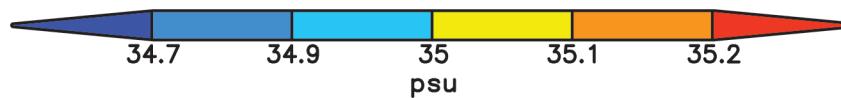
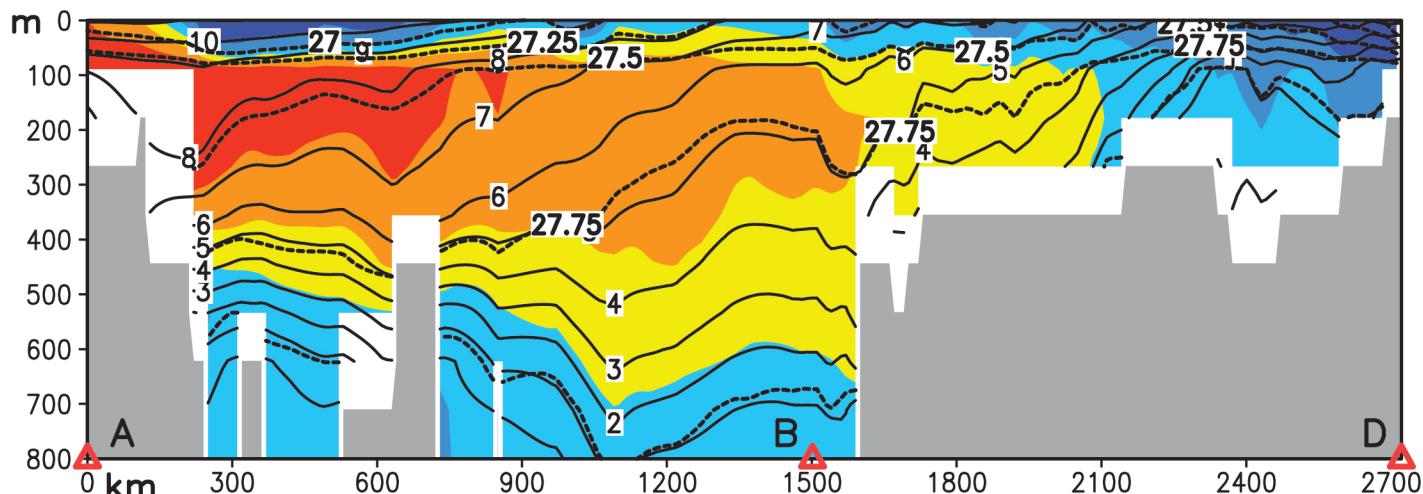
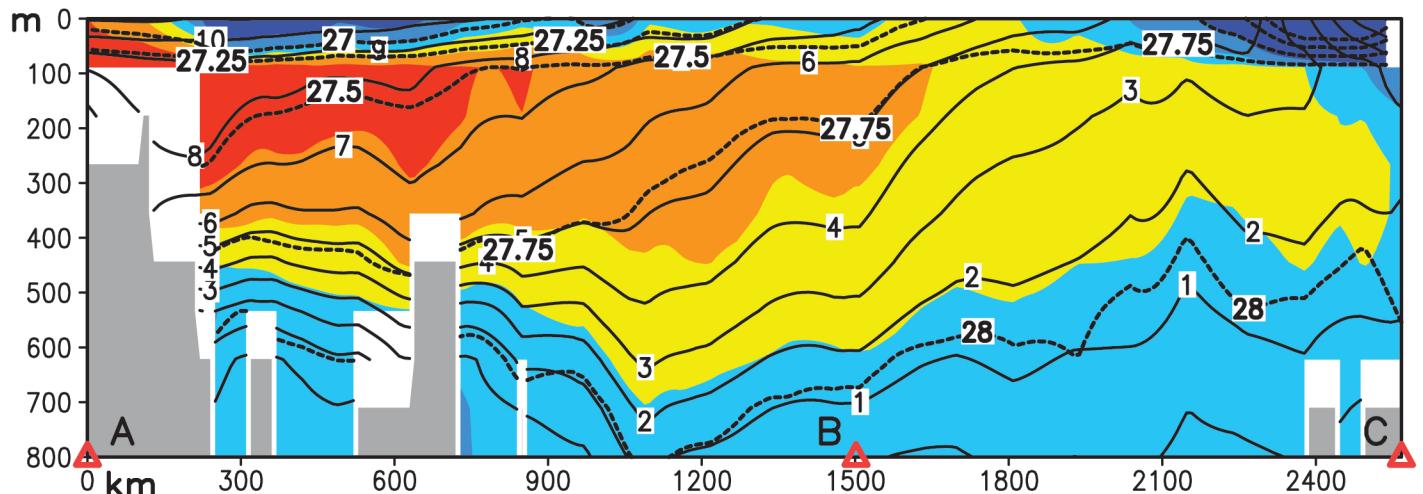
696 **Figure 8** (upper panel) Annual heat content anomalies () integrated
697 throughout the AWZ (where T' is anomalous temperature). Grey line shows the
698 two-year smoothed monthly NAO Index time series obtained from the National
699 Centers for Environmental Prediction (vertical range: ± 0.8). (middle panel)
700 Biannually smoothed rate of heat storage anomalies ($\partial HC / \partial t$, black) and four
701 estimates of net surface heat flux averaged over the AWZ: NCAR/NCEP
702 reanalysis (blue), ERA-40 (red), ERA-Interim (yellow), and WHOI (green).
703 Downward heat flux into the ocean is positive. (Lower panel) Convergence of
704 horizontal heat transport, estimated as the difference between net surface heat flux
705 and rate of heat storage, for each surface heat flux estimate.
706

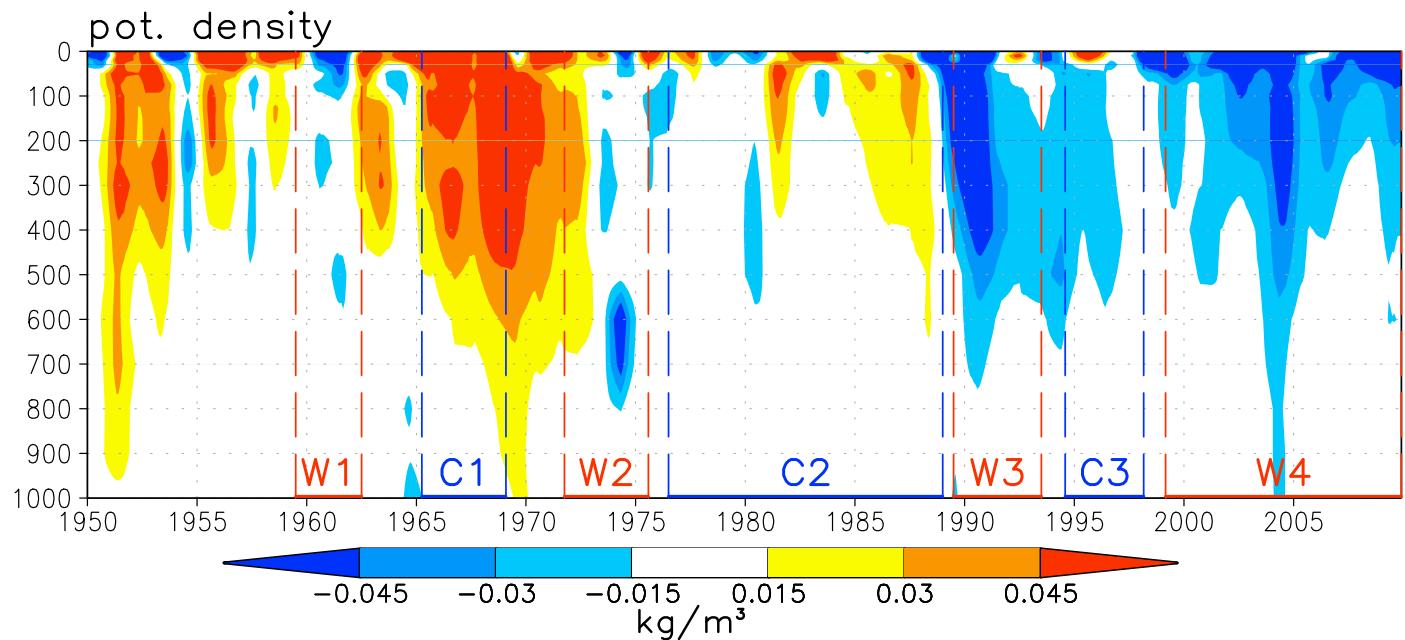
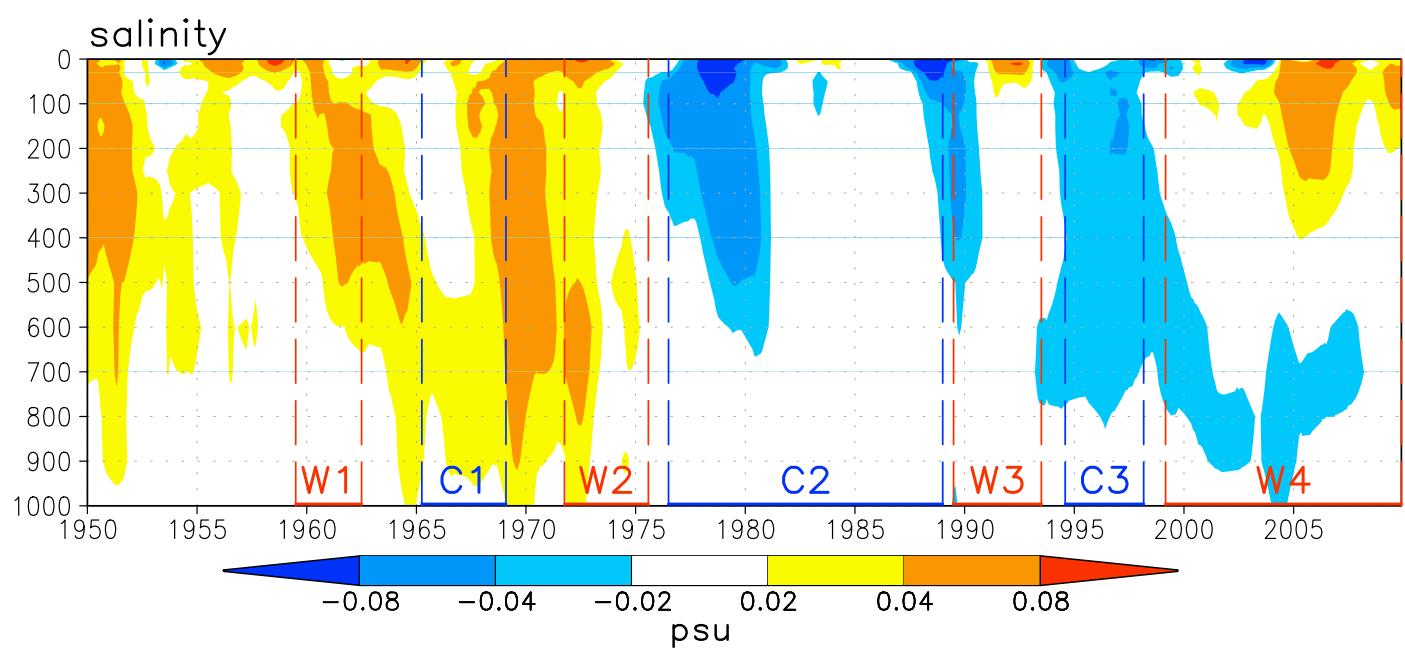
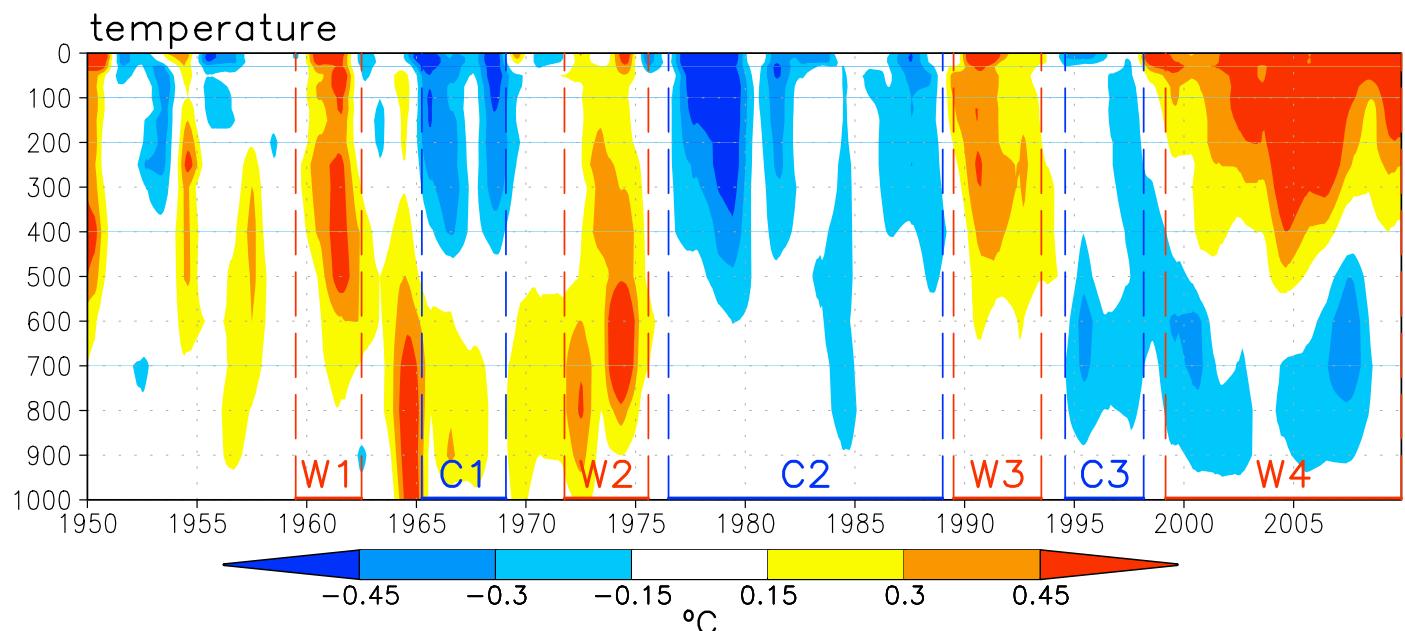




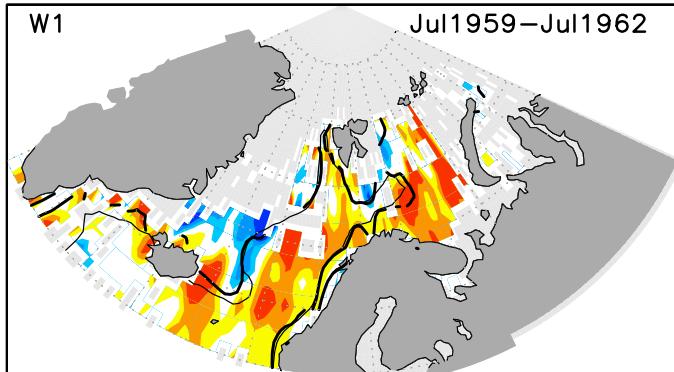
1950–2009, 100m



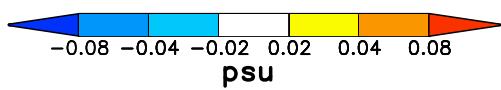
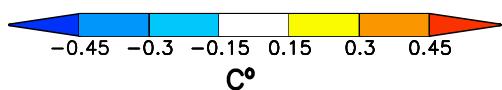
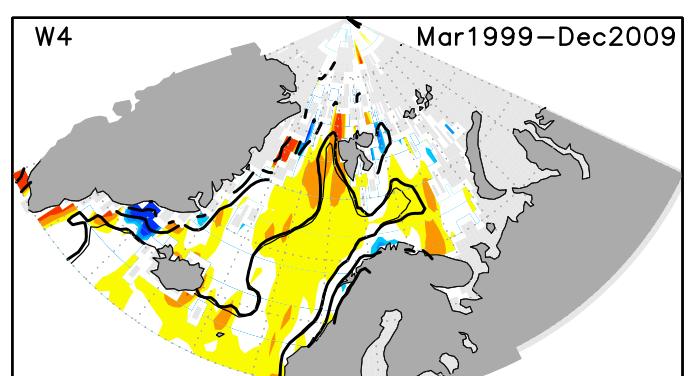
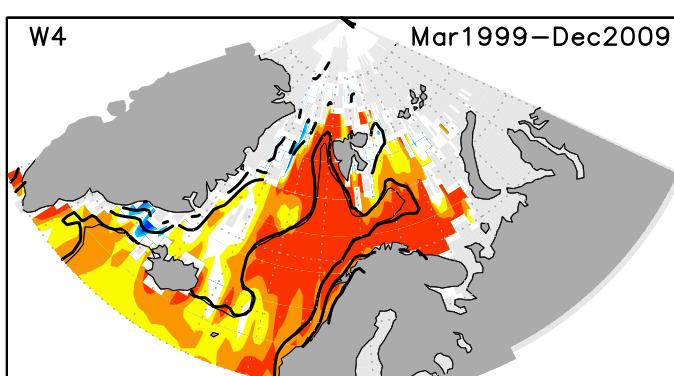
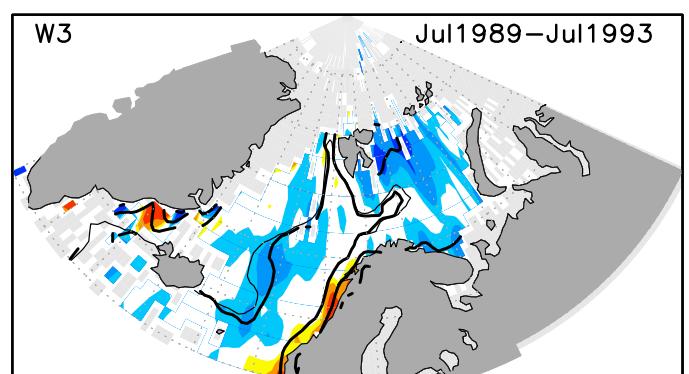
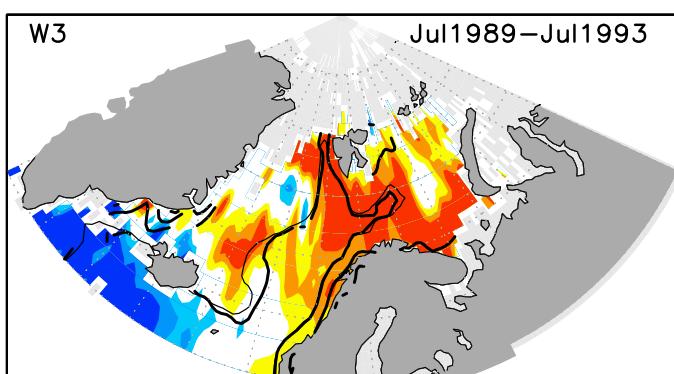
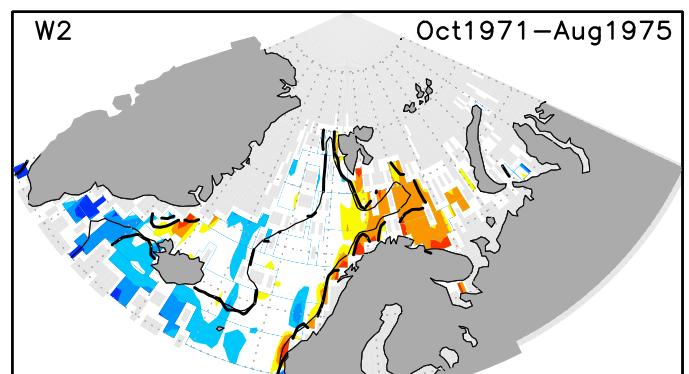
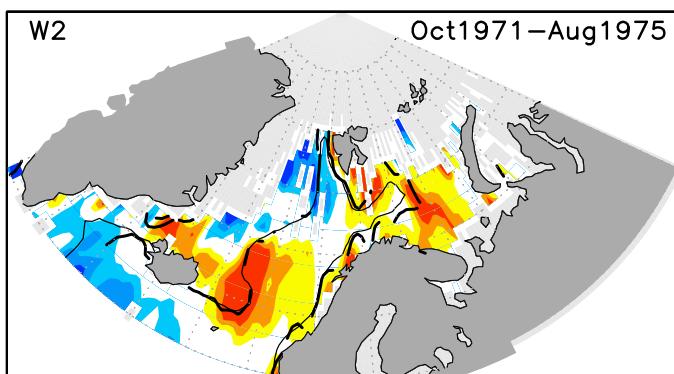
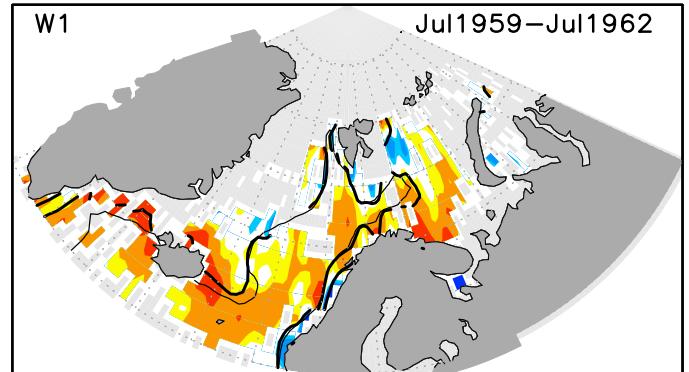




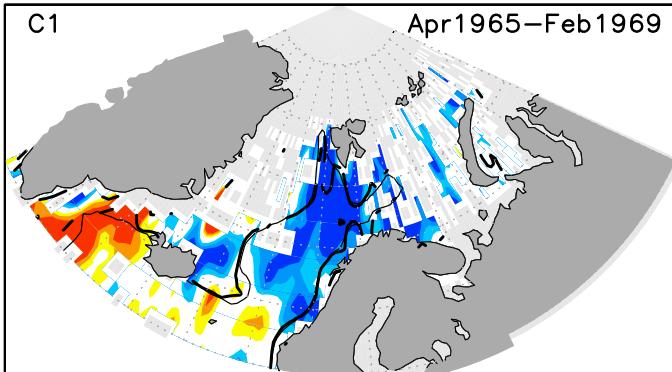
Temperature anomalies



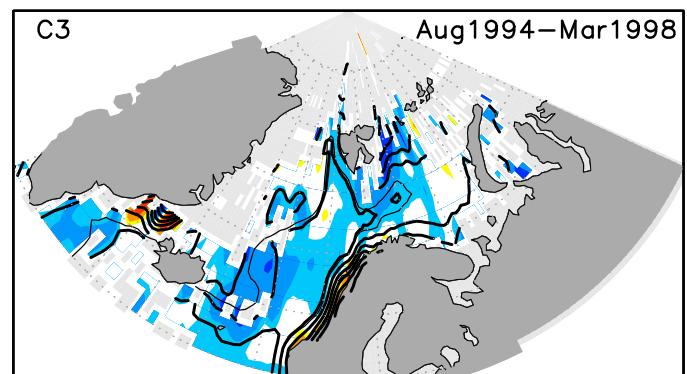
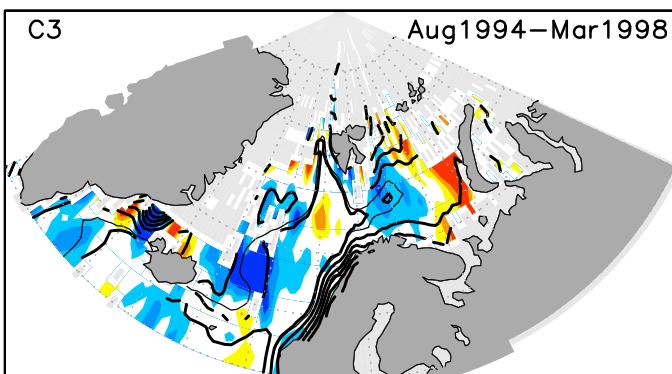
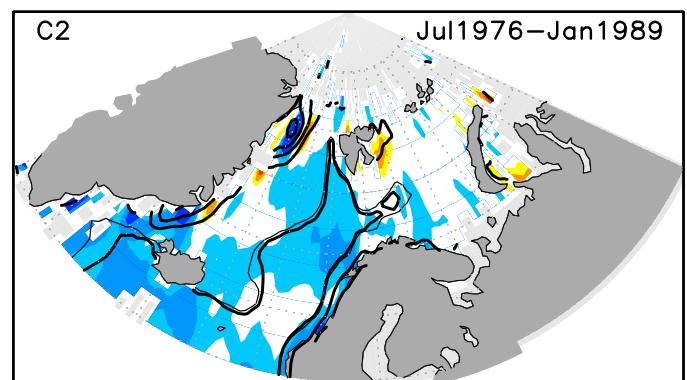
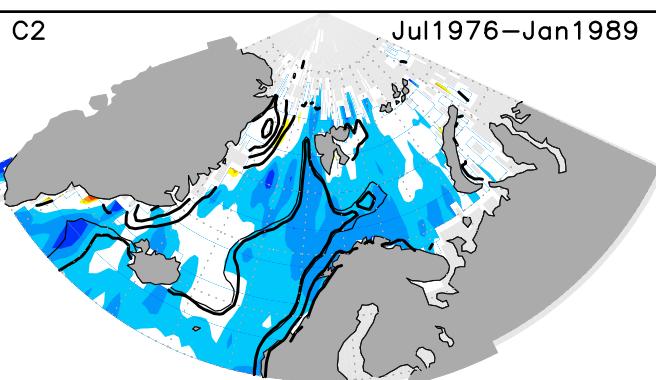
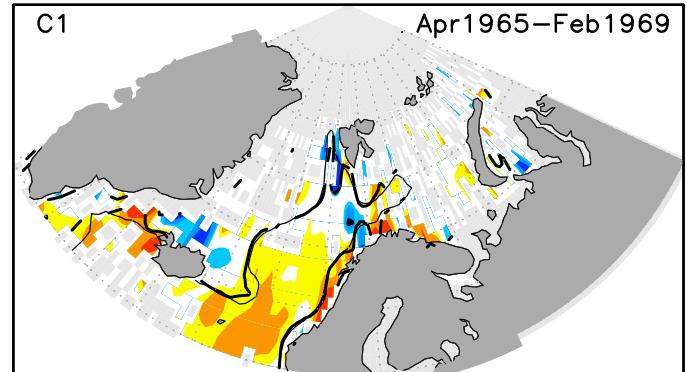
Salinity anomalies



Temperature anomalies

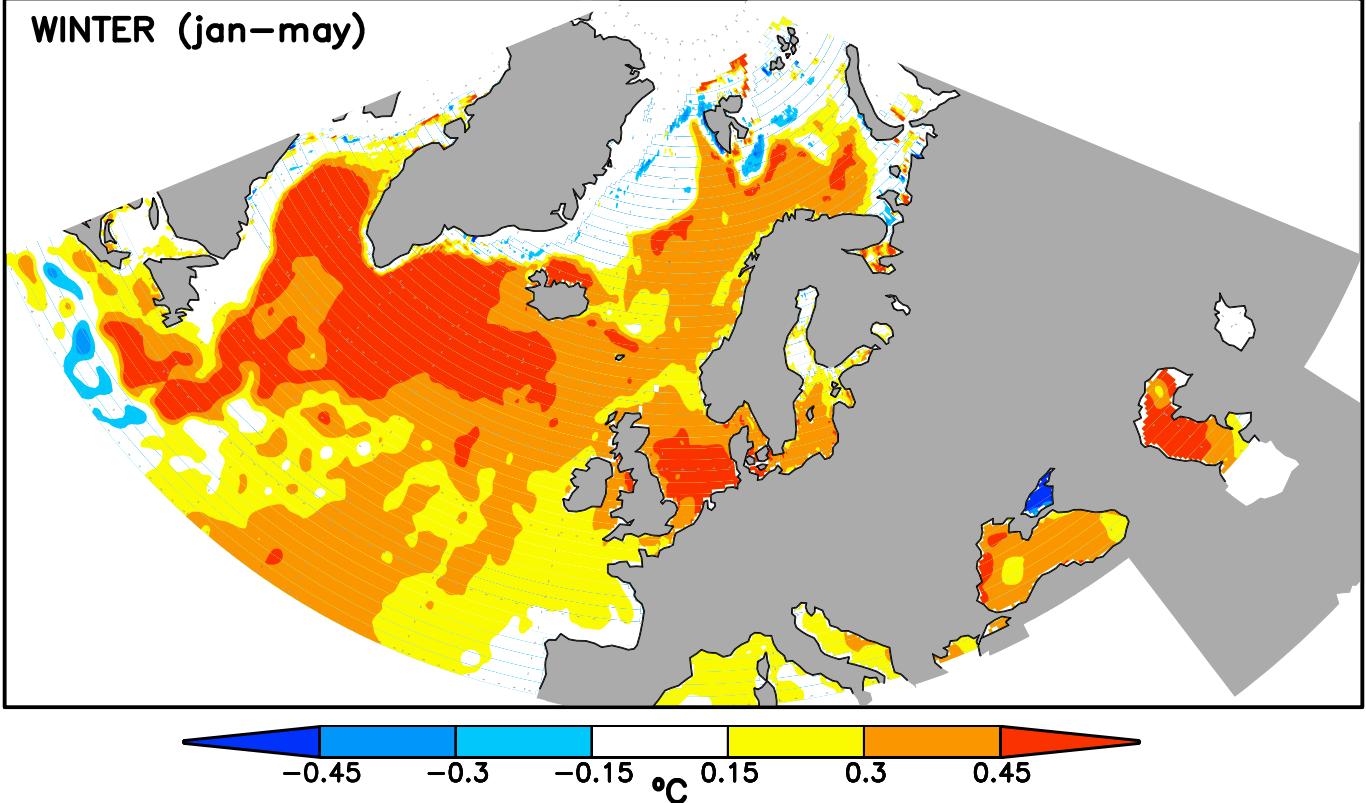


Salinity anomalies



SST anomaly (Mar1999–Dec2010)

WINTER (jan–may)



SUMMER (jul–oct)

